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SAND WAVE STUDY

by
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This report will provide further information on the origins of depositional bed forms and their attendant flow regimes. It will focus on sand wave formation, size, and dynamics because the occurence of these features can have far reaching implications on site selection efforts and the integrity of buried objects such as cables and pipelines.

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1. INTRODUCTION

Ocean engineering and navigational projects increasingly require such information as sea bed shape and stability for accurate designs, so it is imperative that the complex mechanisms involved in the formation of erosional and depositional bed forms be thoroughly understood. In doing so, scientists and engineers from various disciplines have investigated the origins of those sedimentary features formed at the boundary between flowing water and a mobile bed of unconsolidated sediment. Attention has been devoted to the study of sand wave fields because it is here that the sediments are particularly prone to dynamic change and can adversely impact foundation engineering designs.

1.1 Scope of Study

As an extension of a previous summary paper entitled, "Bed Forms: Geological Effects of Bottom Currents," this report will provide further information on the origins of depositional bed forms and their attendant flow regimes (De Visser 1996). It will focus on sand wave formation, size, and dynamics because the occurrence of these features can have far reaching implications on site selection efforts and the integrity of buried objects such as cables and pipelines.

1.2 Background

Erosional features like channels and troughs tend to form along the axis of deep-flowing currents while depositional features tend to appear on the edges of these currents. The largest amount of erosion and, consequently, the largest erosional features are commonly found in such bottom-water pathways as ocean gateways and interbasinal channels. An area with an absence of mounds, tracks, or burrows may at first glance seem devoid of sedimentary features when in fact it is a subtle indication of the most gentle of erosional features. For example, poorly formed sedimentary features are a result of several factors: the periodicity of ebb and flow tides, the irregularity of non-tidal currents and waves, the more extreme variations in depth of flow for rivers and tidal areas than on the ocean continental shelf, and the more consistent sand replenishment that takes place in rivers and experimental flumes than in oceans. Small-scale features can be concentrated in an area where the flow velocities are localized because chemical dissolution and mechanical erosion processes are more pronounced in this area than in surrounding areas that are more protected. The growth of larger features such as seamounts, ridges, escarpments and other unconformities is then also directly related to the streamlining and accelerating water masses that flow around these obstructions (Kennett 1982).

Depositional features created by bottom currents are found throughout the ocean and bear a striking resemblance to the aeolian sand dunes found in deserts, though the aeolian features can become much larger than their aqueous counterparts. According to Kennett (1982), the erosion and deposition cycle promotes the formation of features which range from small, temporal lineations and ripples centimeters in size to sediment drifts several kilometers in crest-to-crest length which have taken hundreds of thousands of years to form (Figure 1).

The western side of ocean basins are known to exhibit the most dramatic bed forms and it is in the Atlantic Ocean, with its large amounts of glacially-derived terriginous material, that prominent abyssal bed forms are found. This may be attributed to the greater dissolution of sediment involved in the bottom sediment transport that takes place in the Pacific (Kennett 1982).

The cycle of erosion, transportation, and deposition by bottom currents in the ocean depends on several factors ranging from the rate at which sediment is supplied and removed from the sea floor, to water flow rates, direction of flow, and sediment characteristics. Sediment supply rates are controlled by biogenic productivity and terriginous input whereas the rate of sediment removal is influenced by bottom flow rates and the dissolution of biogenic sediments by bottom currents. For example, sediment thickness can vary over short distances, but since the pelagic material supplied to the ocean floor is constant over large distances, it can be assumed that any topographic irregularities are attributable to the influence of bottom currents.

Besides the rate of supply of sediments to a site, the specific nature of an individual bed form depends on water flow rates and direction of flow. Slow rates of water flow cause the formation of features oriented either perpendicular to currents, as in the case of ripples, or those situated within approximately 20° of parallel to the local peak flow direction as is demonstrated by sand ribbons and tidal current ridges. As the flow speed increases, sedimentary structures form in a direction normal to the flow direction and create megaripples and sand waves. Asymmetry in the bed form's cross section can arise when the net flow is in one direction, a characteristic that is evident in the gentle slopes that form on the upstream side of a feature and the steeper face on the downstream side. Steady flow oscillations encourage symmetry in the cross section and as flow rates increase, parallel features such as banks, develop (Kennett 1982). Ocean currents can exhibit the combined influences of tidal energy dissipation and the net transport of bottom water from one region to another. The periodicity of tidal currents also affects the growth of depositional features in the deep ocean, but it is an aspect of the dynamics which is poorly understood, though their role in shallow regions is very important in the formation of small scale sedimentary features (Kennett 1982). The morphodynamics of a bed form also depends on the characteristics of the sediment, since it appears that large supplies of noncohesive sands in a particular size range are necessary in order for sand wave formation to take place (Levin 1992).

Though bed forms can be easily identified when they are well developed, they often exist in nature as complicated combinations of several shapes because of poor formation or sparse distributions which makes them more difficult to recognize. The lowest-order form in a bed form classification scheme is generally considered to be the ripple which develops at a water velocity just above the initiation of sediment motion. Sleath's schematic of the principal bed forms produced by steady currents in the sea is shown in Figure 2; typical bed form heights are given as a function of either the median grain size of the sediment, D, or in meters, m (1984). Note that the very small scale feature known as "parting lineation" in Figure 2 is generally ignored in bed roughness or sediment transport calculations, so it is not usually considered to be the lowest order form in many classification systems. As flow velocities increase, the development of the next size class of bed forms occurs. It is important to realize that various bed form classification schemes exist, each of which varies with respect to the definitions of their constituent bed forms and associated boundaries. For example, Blatt et al.'s (1980) table of bed form characteristics shows representative values for bed form length (wavelength), vertical form index (wavelength-to-wave height ratio), and grain size, provides geometry descriptions, and defines the controlling hydraulic variables for these features (Table 1). Kennett (1982) also gives descriptions for very small ripples with spacings less than 60 cm and heights ranging from being barely perceptible to 20 cm or more, sand waves or dunes with typical lateral dimensions of 10 to 100 m, and depositional ridges hundreds of kilometers long and tens of kilometers wide (Kennett 1982). Bed form configuration changes slowly with time, but the same general characteristics are preserved over a range of flow and sediment conditions.

Studies conducted in both laboratory and natural environments have provided insight into the area of sand wave development. Since a set of bed forms, or the bed configuration, changes slowly with time, however, the smaller forms are most easily studied in laboratory flumes where conditions can be tightly controlled. Therefore, larger amounts of quantitative data exist for sedimentary features formed by the small, steady unidirectional flows generated in laboratory flumes than data obtained for the larger scale bed forms found in nature. The larger bed forms are tracked in nature where settings range from rivers and sandy coastal embayments to the wide expanse of the continental shelf. The unidirectional and channelized flow characteristic of rivers yields a wide range of grain sizes and flow regimes, while sandy coastal embayments exhibit channelized, unsteady, and tidal flows which ebb and flow. Though intermediate-sized features can be studied in low water river flow or during low tide in estuaries, these conditions are not entirely comparable to the conditions that occur in the ocean. As such, the very deep, unchannelized continental shelf region is dominated by tidal currents, geostrophic flows, storms, and wave-generated currents that can be studied by acoustic and other methods available to researchers (Mahmood 1981; Stride 1982).

2. OCCURRENCES OF SAND WAVES

Sand waves are widely distributed in several types of tidal environments and have been studied by scientists and engineers in the field. These features are found in complex channels of sandy coastal barriers and estuarine and shoal areas, tide-swept seas and seaways, and shelf environments and shallow marine platforms. The features have been intensively studied in a number of estuarine areas: Parker River Estuary in Massachusetts (Daboll 1969) at the entrance of Chesapeake Bay (Ludwick 1972), and in the James River Estuary, Virginia (Nichols 1972). Sand waves have also been found in European estuarine and shoal areas: Gibralter Point, Lincolnshire, England (King 1964), in the mouth of the Tay Estuary in Scotland (Green 1975), in the lower St. Lawrence River of Canada (D'Anglejan 1971), and in the Outer Thames Estuary (Langhorne 1973). Classic examples of riverine sand waves that form and migrate under conditions where flow is typically downstream include the Mississippi River (Wright 1974) and the Columbia River in Oregon (Whetten 1986). Tidal currents have generated two- and three-dimensional sand waves on the floor of central San Francisco Bay

(Rubin 1980) and several researchers have surveyed sand waves in Lower Cook Inlet, Alaska (Bouma 1980; Mahmood 1981).

Also found in restricted, tide-swept seas and seaways, sand waves have been identified in the Malacca and Taiwan Straits (Keller 1967), Japan's Inland Sea (Ozasa 1974), Long Island Sound (Bokuniewicz 1977), and the Bay of Fundy (Swift 1968). An extensive number of studies have been conducted by several scientists in the seas surrounding the British Isles: Stride (1973), McCave (1971), Caston and Stride (1973), Hails and Kelland (1974), and Langhorne (1982). Sand waves have also been identified along several stretches of the shoreline of The Netherlands by Verhagen (1989). Morang and McMaster (1980) detected fields of low amplitude sand waves in the nearshore adjacent to the southwestern Rhode Island beaches. Hunt et al. (1977) studied the sand waves on the shelf environment of Diamond Shoals, North Carolina, and the features were also identified on the shallow marine platforms of Georges Shoals and on Lily Bank, Bahamas, by Stewart and Jordon (1964) and Hine (1977), respectively.

3. SAND WAVES AND ENGINEERING INSTALLATIONS

Quantifying the dynamic conditions responsible for sand wave development can be crucial to the success of maintaining navigable waterways and ensuring the integrity of such objects as pipelines or cables. For example, the existence of sand waves may affect the proposed installation of an underwater sound range offshore of Onslow Bay, North Carolina. Smooth sand areas on the shallower portion of the continental slope merge into an area where large scale sand waves have been identified in deeper waters. The bottom currents stemming from the Gulf scour and erode the area, processes which could alternately expose and bury hydrophone mounts (Paquette 1995; De Alteris 1996).

Buried mines in the approaches to major ports and in shipping choke points also constitute a hazard, especially when sand waves are one of the burial mechanisms in affect. Such locations as Australia's Charles Point Patches in the approaches to Darwin and in various choke points within Torres Strait are identified by Mulhearn (1996) as areas which are impacted by mobile sand dunes. His mathematical model shows that the factors most critical for the time taken for a mine to become buried are primarily current strength, followed by dune (sand wave) size, and lastly, the initial location of a mine in relation to crests and troughs of a sand dune field. The results show that as current strength increases, the time taken for a mine to become buried decreases sharply. However, the duration of time required for burial increases as a dune's size increases and as a mine's initial distance downstream from a dune's crest increases. Though this model was tailored for those cases in which dune heights are much larger than mine diameters, the results remain indicative of the critical factors influencing burial by migrating dunes (Mulhearn 1996).

Additionally, the sand wave is the most important bed form in relation to marine pipelines in that they, as a potential foundation hazard, can sometimes reach large enough sizes and migrate fast enough to cause significant bed changes with respect to the diameter of the pipeline. As was mentioned earlier, sand wave formation is not wholly confined to tidal areas either, and its occurrence in deep water can result from density differences which cause relatively high current flows near the bed. Thus the design of pipelines is substantially influenced by the determination of whether or not trenching or burial is necessary, evaluating the erodibility of the seabed, scouring, free span development, the potential for liquifaction, self-lowering and natural backfilling, lateral and vertical stability, and the response of exposed or suspended pipeline sections to hydrodynamic loads.

The significant impact that these bed forms can have on engineering installations is exemplified by a 36-in. oil-export pipeline that was installed in the North Sea in 1990 (Steel 1991). Sand waves up to 3 m high were found to blanket an area between 5 to 30 km from shore in water between 50 and 100 m deep. Bathymetric irregularities such as these can cause portions of a submarine pipeline to be unsupported by the bed, thereby forming freely spanning sections and areas of significant off-bottom clearance. These, in turn, can cause sections of the pipeline to experience unacceptable bending moments, flow-induced vibrational damage, and buckling instabilities. In the case of the new pipeline that was installed parallel to and in the same corridor as an existing line between the Forties field and Cruden Bay on the United Kingdom mainland, the unexpected presence of sand waves potentially could have caused considerable increases to both project costs and installation scheduling delays.

Steel and Inglis (1991) state that a pipeline operator has two options available when confronted with free spans: either excavate the seabed at the high spot and lower the pipeline to increase the support, or provide intermediate support to the individual spanning sections. The selection of either of these options is a function of the number and severity of the locations requiring remediation, water depth, seabed composition, pipeline parameters, and scheduling. Prominent sand waves were identified at 6, 8, and 10 km from shore and a series of undulating forms in 50 to 100 m of water, depths the authors state were much greater than they would normally expect in regions of sand wave activity.

In order to better quantify conditions at the site, a route profile assessment was conducted. Bottom-roughness analyses performed with various computer programs to define bending-moment distribution along the route centerline. This work indicated that minor route deviations would enable the pipeline to be directed around some of the worst areas, though constraints of the existing pipeline would limit the effectiveness of these adjustments. Nonetheless, the evaluation revealed that (1) the pipeline could be laid and filled with water without prelay rectification work, though the sand wave peak situated 8 km offshore appeared slightly overstressed in this water-filled state, (2) seven other locations required post-lay remedial work, and (3) the acceptability of these and, possibly, additional locations were sensitive to even minor changes in the survey profile as could be expected from the normal installation tolerance of 10 m from the route's centerline and any localized settling that may have taken place at support points (Steel 1991).

An evaluation of potential rectification systems showed the most cost-effective solution for modifying the profile along the new pipeline route would be to use a tracked suction dredger deployed from a dive-support vessel. The span limits were also evaluated by using fatigue limit state criteria and an allowable strain, which showed that the sand wave situated 8 km from the beach would not adversely impact the installation. Therefore, remedial work was postponed till after the pipelay. These evaluations negated the need for $9.5 to $12.5 million of prelay dredging and a lengthy lead time of 6 months to modify the dredgers beyond their normal water depth range (Steel 1991).

Steel and Inglis state that (1991) an interferometric swath bathymetry system was also used in conducting the prelay survey. This system was used to provide a higher density of soundings than a conventional echo sounder, and, in turn, a more detailed view of the sea floor, thereby allowing for more accurate rerouting of the pipeline; it also enabled the post-lay rectification work to be more closely defined. Profiles for several route options revealed the following information: (1) the general features of the pipelay route were consistent with an earlier survey that was conducted, (2) additional route deviations were minimal, and (3) a total of eleven areas would require post-lay remediation.

As-laid survey results showed the major sand wave ridges. Subsequently, a Digiquartz survey of each work site was conducted before dredging units were deployed. Various measurements were then taken by saturation divers and another Digiquartz survey was performed. The revised profile showed one extreme case where 8 km offshore, the pipeline spans on either side of the sand wave ridge were 100 m long, 2 m off the sea floor, and 80 m long, 1.5 m off the sea floor. This ridge was then excavated to remove the spans and allow the pipeline to be trenched to the cover specification of 1 m. Survey and rectification work lead to a total of six sites being dredged and then measured by divers, without affecting the project's overall schedule; this course of action also turned out to be less costly than performing prelay dredging (Steel 1991).

Not only may sand waves and sandy ocean bottoms adversely affect such buried objects as pipelines, their interaction may result in spontaneous natural burial of lines that, if accurately predicted, may enable operators to realize significant cost savings if the extent of mechanical burial is known beforehand. Schaap (1989) states that Protech International developed a natural burial prediction model based on theoretical and experimental research that analyzes the behavior of submarine pipelines in moving sand waves. The model was calibrated using survey data of ten prototype pipelines and it has been successfully applied over a period of seven years prior to pipeline projects. The author indicates that submarine pipelines situated in moving sand wave areas can eventually be buried completely and thereby reduce trenching and dredging requirements if the pipeline's structural strength allows for safe installation, testing, and operation during the burial process (Schaap 1989).

Protech International's computer program was used to analyze the behavior of submarine pipelines under the influence of moving sand waves by taking the non-linear geometrical and material properties of the pipeline into account. The computer program's results are then compared with values for the ultimate deformation capacity of steel tubulars subjected to internal and external pressure. Burial requirements for submarine pipelines on the Dutch Continental Shelf pipeline network, 25% (217 km) of which has been subjected to natural burial processes, are related to risk analyses. Criteria are necessary to prevent oil spills and such external hazards as the destruction of the pipeline by bottom fishing gear, the most frequent cause of damage to submarine pipelines (Schaap 1989).

Data from the first ten of twenty pipelines in the Dutch Continental Shelf pipeline network subjected to natural burial form the basis of the successful empirical natural burial prediction model described by Schaap (1989). This model shows the predicted behavior to be in good agreement with observed behavior. It also shows that the pipeline disrupts the natural flow regime surrounding the pipeline, thereby causing increased erosion near it. Pipelines undergoing natural burial experience two types of erosion: lee-erosion and possibly concurrent tunnel erosion. Tunnel erosion is a short-term process which acts on the order of hours or days; as scouring processes occur due to sea floor irregularities or water movement, a tunnel forms underneath the pipeline. Lee-erosion, on the other hand, is a process whose time scale is on the order of weeks or months; it also affects a larger area than tunnel erosion. A pipeline on the sea floor disturbs the flow around it, causing turbulence on the pipe's lee side which, in turn, allows sand to be transported.

According to Schaap, strong, long-lasting currents induce the lee and tunnel erosion phenomena to naturally lower the pipeline into the scour hole due to the pipeline's submerged weight and flexibility. This lowering process stops when bed load transport no longer occurs and when variable flow conditions result in the natural backfilling of the eroded pipeline trench; these processes can also occur simultaneously (Schaap 1989).

As an alternative to the costly removal of sand waves from the pipeline route before installation begins, Schaap (1989) suggests the peak shaving of sand waves as a cheaper option, one that can be followed by trenching in order to use the deformation capacity of the pipeline to follow the sea floor's contours. However, only elastic bending or limited plastic deformation is acceptable, as is dictated by international pipeline codes. Increasing or reducing the tensioning during pipelaying or dredging also influences the pipeline response, though the economics of the installation will generally govern the choice of measures taken to ensure a successful pipeline installation and operation.

Self-lowering of a pipeline on an erodible seabed in shallow waters has also been investigated by Bruschi, Sintini, Johannesen, and Verley (1993). A portion of the Sleipner to Zeebrugge pipeline on the Dutch Continental Shelf was to be installed with a 40 in. OD diameter pipe in shallow waters. More than 350 km of pipeline were to be installed in water depths less than 35 m, in an area affected by severe environmental conditions, sand waves, and such activities as fishing and ship traffic. In order to generate an appropriate installation design, Bruschi et al. (1993) state that it is necessary to evaluate the erodibility of the seabed, scouring action, the potential for soil liquifaction, free-span development, and self-lowering. If on-bottom stability criteria are met, these authors state that a decision regarding whether or not to perform post-trenching of an area can be made after evaluating the risks to the pipeline during the first winter season and studying survey results taken early in the second season.

In addition to the erosion and deposition phenomena resulting from modification of the flow field surrounding the pipeline area, Bruschi et al. (1993) state that pipeline lowering may be facilitated by partial collapse of the support at the span shoulders, caused by the longitudinal propagation of erosion, or wave-induced liquefaction, combined with the pipeline's submerged weight. The characteristics of the seabed affect the permeability and, therefore, the liquefaction and seepage pattern underneath the pipeline. Consequently, the stratigraphy of the seabed may limit the depth of erosion and, thus, the capacity for self-lowering. These various phenomena can cause the pipeline to become partially lowered below the seabed for long periods, partially free spanning, or still rest on an undisturbed bed.

Bruschi et al. (1993) used models by Bryndum et al. (1991) and Bernetti et al. (1990) to predict the self-lowering of the Zeepipe. Bryndum et al.'s (1991) two-dimensional model takes into account tunnel erosion and leeside erosion, and is based on the fact that after having attained immediate lowering due to tunnel erosion, the pipe will still protrude from the bed. Consequently, the leeside erosion process acts to lower the seabed around the pipe, which then allows further tunnel erosion to take place. As the pipeline and surrounding bed sink with respect to the original seabed, the eroded hole increases its capacity to trap transported sediment. In this manner, the backfilling process will dominate over the erosion process, eventually either burying or partially burying the pipeline.

Bernetti et al.'s (1990) three-dimensional model is based on simplified hydrodynamic and sediment transport formulations, but also includes pipeline flexibility. The time scale for self-lowering in this model is evaluated by basing it on the development of scour-induced free spans as the early stage of the lowering process. Bruschi et al. (1993) used each of these models to assess the potential for self-lowering. The results of the analyses allowed for design criteria to be developed for different sections of the pipeline with respect to the environmental conditions in the area of consideration. Thereafter a monitoring program was initiated after the first lay-season to obtain data related to free-span development and the self-lowering process.

A self-burial pipeline survey was conducted after the first winter season and the self-lowering expectations were compared with the in-field observations. Thus, areas were identified where post-trenching was required prior to the pipeline's operating phase. It was determined that the amount of self-lowering and the time scale for this phenomenon to occur in medium to dense sand areas are comparable with the predictions made with Bryndum et al.'s (Bryndum 1991) model. Furthermore, it was found that in areas of loose sediment overlaying clay or dense layers of sand, the susceptibility of the sand to liquefaction processes would make it difficult to accurately predict the extent of lowering. Nonetheless, as a result of immediate settlement and the erosion process, a minimum lowering of approximately 0.5 D (external pipeline diameter) in these areas was assumed. The evaluation indicated that seven areas required further trenching and, in particular, the observed time-scale of self-lowering and the final configuration of the pipeline with respect to the location of the original sea floor suggest that tunnel erosion is the major mechanism in causing self-lowering in sand-filled areas affected by cyclic pressures induced by waves (Bruschi 1993).

Many other computer models exist for determining sand wave heights and migration rates and calculating pipeline/cable seabed interactions. Staub and Bijker's (1990) numerical models are representative of these in that their simulations require comprehensive environmental input data in the form of time series of wave and current data. Their sand wave model modifies and extends existing theory to account for non-equilibrium sand waves, varying directions of waves and currents, and the effects of short-crested sand waves, high water waves, and wave-generated ripples. Their self-burial model integrates the effects of the physical processes of tunnel erosion, leeside erosion, pipe lowering and backfilling, and is based on flume tests and numerical models.

Staub and Bijker's (1990) self-burial model is based on observations of tunnel erosion and leeside erosion; these authors also indicate that flume tests and field data show backfilling to be an important factor in pipeline burial. During the burial process itself, the pipeline and the seafloor around it begin to sink relative to the "original" position of the seafloor. Consequently, the upstream eroded hole begins to increase its capacity to capture sediment which, in turn, allows the backfilling process to become increasingly dominant over the process of leeside erosion. Staub and Bijker (1990) indicate that such environmental conditions as waves and currents primarily determine the equilibrium position of the pipeline; taking this one step further, each individual set of environmental conditions causes a different equilibrium position. Thus, one set of environmental conditions, such as those that constitute a storm, may cause a pipeline to diverge from its previous equilibrium position. The effects of pipeline stiffness is not included in this two-dimensional model, but is taken into account by its calibration against actual three-dimensional pipelines.

As is the very nature, numerical models are constantly being revised and updated, and future extensions of existing models will improve upon previous work by taking into account additional factors.

4. SITE ASSESSMENT

Various methods are available for investigating seafloor geological conditions, especially sand waves. Such high-resolution geophysical tools and methods as acoustipulse, fathometer, and side-scan sonar are usually supplemented with bottom sampling, soil borings, and visual reconnaissance (i.e., bottom television and camera); a majority of these techniques were used by Mahmood, Ehlers, and Cilweck (1981) to investigate sand waves in Lower Cook Inlet, Alaska. The acoustipulse system is a single channel, high resolution, marine reflection seismic system developed for high resolution and penetration of up to 90 m in soil and soft-rock strata. This electromechanical acoustic profiling system's pulse typically penetrates the seafloor to approximately 70 to 90 m, though penetrations of up to 150 m can be obtained in soft materials. Reflected signals are then received by a hydrophone array and the data displaced graphically with typical vertical exaggerations of approximately 10 to 20 times. The fathometer is used to transmit sound energy downward to the seafloor and its reflection back to the transducer is recorded and converted to a graphical recording; thus the amount of time lapsed between the pulse transmission and the echo reception is a measure of the distance traveled through the water column, a value that is then converted to depth. The side-scan sonar provides a graphical record that shows the two-dimensional nature of the sea floor. Features recorded with this instrument commonly appear similar to their natural perspective and topographic irregularities are discernible. Radar imaging in coastal water with strong tidal currents has also been successful, as is exemplified by the revealing sea bottom topography identified in a Seasat synthetic aperture radar (SAR) image of the Sandettie sand bank in the North Sea (Hennings 1990). The digitally processed radar image allows bathymetric radar signatures, shown as alternating black and white bands, to be identified.

Even though various techniques are available for assessing a site, an integrated, multi-scale observation and sampling program is necessary in order to accurately characterize bottom topography. Bouma, Rappeport, Orlando, and Hampton (1980) compared data from high-resolution seismic profiling systems, side-scan sonar, television, camera, and bottom sampling and determined that bottom and bed form interpretations based solely on sonographs can be inaccurate. They found the length of acoustic shadows from sonographs for obtaining bed form height was exaggerated by 3 to 7 times. Television feedback showed the troughs between small sand waves to be flat and filled with shell fragments; this coarse material has a high acoustic reflectance that can be easily misinterpreted with a similar acoustic reflectance received from the slope of a sandy bottom. Therefore, if a researcher seeks to calculate bed form height by relying solely on side-scan sonar records of bed forms that are not also visible on bathymetric records, erroneous calculations would result if these data are not supplemented with photographic records.

Since sand wave morphology can be grossly exaggerated in side-scan sonar and echo-sounder records, some researchers choose to use more conventional field techniques to assess a site, though such measurement programs may be tedious, expensive, and difficult to conduct. Langhorne (1982) studied the dynamics of a sand wave by using cross-sectional profiles obtained from lines of sea-bed reference stakes. The measurement scheme consisted of divers hammering a line of mild steel reference stakes into the seafloor at the beginning of the experiment. The stakes were introduced at 50 cm intervals that crossed the sand wave's crest at right angles; then each stake was driven to a level height of reference with respect to its neighbors. To compensate for the sloping areas of a sand wave, the stake heights were incrementally varied along the line of stakes. The tops were also leveled with respect to a 350 kg sinker module which acted as a local bench mark from which the stake location could be determined if erosion caused the stake location to shift. After the stakes were set up in this manner, divers measured the lengths of exposed stakes to obtain the cross-sectional profile of the crest of a sand wave and changes in that profile. The measurements were obtained over a period of six months to determine sand wave changes in response to tidal and wave conditions; he also used fluorescent tracers to study sediment dispersion from the sand wave crest. The methods used by Langhorne may not be as elegant as the high resolution acoustics mentioned above, but they overcome many of the accuracy and repeatability problems experienced by shipborne survey techniques.

5. FLUME BED FORMS

Though bed form research efforts date back to the early 1900s with the pioneering studies of Cornish (1901), Bucher (1919), and Van Veen (1935), no definitive relationships exist for sand wave dynamics. Nonetheless, the large amounts of quantitative flume data that exist can provide useful insights into the behavior of larger bed forms in the field if the appropriate boundary conditions are taken into account. For example, if phenomenon are specific only to small depths, the effect cannot be extrapolated to large depths. To take into account the fully developed turbulent conditions that a bed form would experience in the field, data are taken at sufficiently large distances from the flume entrance, walls, and exit. But before delving into the simplified relationships for these artificially generated features, it is necessary for the reader to understand the nomenclature associated with bed forms.

5.1 Bed Form Mechanics

The nomenclature associated with a subaqueous bed form's shape and components is shown in Figure 3. The bed form wavelength, L, is the distance between troughs or crests measured parallel with the ripple base and perpendicular to the crest. The sand wave height, H, is the interval between the trough and crest perpendicular to the base; the wave amplitude is half its height, H/2, and equal to the amplitude, A. Zarillo (1982) states that it is possible for the wavelength to be up to one hundred times greater than the wave amplitude but the ratio of wavelength to wave height is usually 30:1 and normally not less than 10:1. In the case of an asymmetric bed form, which is indicative of the direction of net sediment transport, the gently sloping side situated between the trough and crest is its "stoss" or upstream side and the steeper side is the "lee" side; the lee side always faces in the direction of water flow and is, therefore, downstream of the crest. Levin (1992) states that the slope of a sand wave's stoss side is approximately 4° and the lee side's slope is less than or equal to 12°, but may equal that of the angle of repose, the angle of tilt beyond which the sediment becomes unstable and slumps; for sand grains, the angle of repose equals 33° (Levin 1992). With high enough flow velocities, sediment transport begins with the individual grains being pushed up the stoss side of the wave form toward the crest. The grains then tumble down the steep lee side and into the trough of the next sand wave where they are once again available for transport (Terwindt 1971).

Researchers find laboratory flumes to be particularly useful for experiments because bed form investigations can be carried out under steady and uniform unidirectional flow; the flow strength is considered to be steady when flow strength does not vary over time and uniform if it does not vary along the channel section from which measurements are obtained. This data can then be used to assess sediment accumulations in natural environments. The laboratory flume is a large, rectangular tank with glass-sided viewing windows that facilitate the monitoring of an experiment at close range. Ideal flow conditions are modeled by using a pump to recirculate a fixed volume of fluid, usually water, through the flume at a calibrated rate. Such initial boundary conditions as the fluid flow rate through the flume can be changed by either increasing the pump's flow rate or by increasing the flume's tilt to increase gravitational forces. The flow depth can be altered by adding or reducing the volume of water in the flume and fluid density can be varied by changing the temperature of the water or by introducing a fluid of a different density into the tank. Sediment size can also be controlled by the researcher. The effect of each of these variables can be assessed by changing one parameter and holding the others constant.

As water flows horizontally in a flume, the deformation of this viscous fluid yields normal and tangential stresses at all points within the zone of deformation as shown in Figure 4 (Levin 1992). The normal stress of the water acts vertically downward on the bottom of the flume and is measured in terms of pressure intensity, P. The tangential stresses caused by the water moving against the flume bottom are measured in terms of shear intensity, τ. Reynolds (1965) conducted flume experiments to investigate the dynamics of water by using dye tracers to track water molecule movement. His first tests were run with a smooth bottom, without sediment. The dye tracer injected into a slowly moving fluid showed clearly defined threads extending through the length of the flume, an indicator that the streamlines are linear and the flow laminar, or not turbulent. With increased flow velocity, however, the colored thread became mixed with the fluid and its sharp outline became blurred. This irregular mixing pattern, a result of turbulent flow, is a condition that is necessary for sediment transport.

Besides allowing the flow to become steady and uniform, the dynamics of an erodible bed in the laboratory can be accomplished by adding sediment to a flume. The bed is allowed to reach equilibrium with the flow so that there is no change in bed morphology, a state which may take from minutes to days of time to attain. Laboratory flume dimensions are constrained by size and design limitations, so they can only be used to study ripples, the smaller-size bed forms; size restrictions prevent the study of sand waves in a flume because these features are depth constrained and do not generally form in water depths less than 5 meters. Sand waves are therefore better suited to being mathematically modeled and characterized in the field (Boothroyd 1985).

The fluid lift and drag forces produced by turbulence, which in turn promote erosion of sediment from a bed, must exceed the stabilizing forces of the sediment. Figure 5 shows the velocity distribution in the boundary layer at the bed, with the current speed decreasing with depth in the flow field and the shear strength profile increasing downwards into the bed. The main stabilizing force for noncohesive sediments is the immersed particle weight, while cohesive sediments that are of negligible weight are stabilized primarily by interparticle adhesion and organic binding. Cohesive sediments are normally soft and layered throughout with their density and shear strength increasing downwards within the bed. Sediment begins to move when the shear stress, τ, transferred to the bed by the ambient fluid reaches the threshold or "critical" shear stress. The stabilizing forces are overcome when this critical velocity is exceeded and the shear imparted on the substrate is greater than the friction force produced by the weight of the sand; erosion takes place at this moment, the threshold of movement. Even though the threshold of movement is proportional to grain size, one may surmise that the smaller the grain size, the easier it is to transport. However, this assumption holds only when the grain is already in suspension. If grain movement is to be initialized, a smaller-sized grain is not easier to transport than a larger-sized grain; for example, cohesive or "muddy" sediments are small in size, approximately less than 125µ, and their large surface area to mass ratio makes them more difficult to erode than sands (Hjulstrom 1939). Generally speaking, the erodibility of sand is a function of shear strength and grain size as shown Figure 6. This graph shows that the erosion of fine sands requires less bed shear stress than silts and coarse sands do (Nichols 1985).

Sediment is moved by one of two basic transport modes, suspended load or bed load, both of which are influenced by the energy transmitted to the individual grains and the bed itself (Figure 7). Bed load is that part of the sediment load that remains more or less continuously in rolling or sliding contact with the bed due to the shear of water moving above the sediment bed. In suspended load transport, the particles of sediment have been lifted from the bed by turbulence and can fall freely relative to the moving fluid (Nichols 1985). The transport mode is influenced by the flow and bed characteristics at a particular site. Both modes of transport can exist at the same time, so it is hard to predict where bed-load transport ends and suspended-load transport begins. For example, ripples and sand waves modify the near-bottom flow field and bring sediment into suspension, but under turbulent flow conditions, the suspended particles are not necessarily kept in suspension because there may be a continuous exchange of sediment between the suspension mode and the bed (Dyer 1972). Just after flow in a flume reaches velocities of 20 cm/sec, fine sand movement will be initialized and ripples, the smallest of bed forms, will grow as indicated in the bed form growth sequence shown in Figure 8 (Levin 1992).

The mechanics of water flow over a sedimentary structures provides an understanding of bed form migration rates and their external morphology. The shape of an individual feature can also be an indicator of internal cross-stratification and predict the flow strength that produced a given bed form. Researchers have demonstrated that ripples propagate rapidly downstream of any unconformity on the bed due to flow separation over the surface anomaly. Erosion takes place where the flow rejoins the bed and scour and accumulation of sediment occur. As the flow moves along the stoss side of a feature, the current streamlines are in line with the bed form near the crest and grains are pushed up the side of the ripple as shown in Figure 9A (Levin 1992). The flow separates from the bed at the brink of the crest and the finer grains are carried downstream in the jet as suspended load, while the coarser grains fall down the lee side of the bed form as shown in Figure 9B. The flow then reattaches in the down stream trough and causes scour, but trapped in the toe of the slip face are the grains which are incorporated into the bed form till the ripple's crest migrates past (Figure 10). The grains are available for re-transport when they are once again exposed in the trough of the ripple. This continuous process of grains sliding, falling, and settling results in the migration of a train of bed forms along a sandy bed (Boothroyd 1985).

The interdependence of grain size and mean current velocity is illustrative of the hydrodynamic forces at work on transported grains. Postma's modification of Hjulstrom's stream data in Figure 11 shows erosion, transportation, and deposition regimes for current velocity versus grain size (Hjulstrom 1939; Postma 1967). Though the data therein are not directly applicable to the sediment dynamics that takes place in coastal and ocean waters, it serves as a good first approximation of the processes that act there. The curves indicate the critical velocities required for erosion and deposition of a particular grain size. The erosion curve shows that for grains larger than 125μ (fine sand), the erosion velocity increases with grain size; thus, coarser grains require more force to initiate movement than finer ones. Unconsolidated clays and silts require velocities of between 10 and 20 cm/sec while sands require velocities of 20 to 40 cm/sec; however, erosion velocities are also high for consolidated silt and clay. It is also noted from the graph that more force is required to move consolidated clay and silt than fine sand. Once grains are set in motion, very low current velocities are required to maintain finer sediments in suspension. (Hjulstrom 1939; Postma 1967; Kennett 1982).

5.2 Bed Form Classification

The shape and origin of bed forms has been studied by engineers, physical geographers, geologists, and oceanographers. Unfortunately, these researchers do not all agree upon the terms and definitions used to describe these phenomenon, nor even come to a consensus within each of their respective disciplines, as to the appropriate nomenclature. The mixed variety of terms and definitions, while based on bed morphology and reflecting the varying effects of channelization, fluctuating water levels, and unsteady and reversing flows, is also confusing because the nomenclature overlaps, duplicates, and conflicts with itself. In an attempt to resolve these issues, two professional groups each tried to clarify the terminology used by researchers. In 1966, the American Society of Civil Engineers' (ASCE) convened a "Task Force on Bed Forms in Alluvial Channels" to discuss these naming inconsistencies (A.S.C.E. 1966; Ashley 1990). They noted various duplications and overlaps in describing the various shapes and origins of these erosional features (Table 2). The Task Force also recommended that a basic classification system based on the division of bed forms into lower and upper regimes separated by a transition zone be used, one put forth earlier by Simons and Richardson in 1963 (see Figure 12; (Simons 1963; Stride 1982)).

With nomenclature discrepancies still unresolved, a Mid-Year Meeting of the Society of Economic Paleontologist and Mineralogists (SEPM) was assembled in 1987 to re-examine the "Classification of Large-Scale Flow-Transverse Bedforms" (Ashley 1990). This group of scientists came to the consensus that even though a wide variety of bed forms exists in fluvial, coastal, and marine environments, all large-scale transverse bed forms, except for antidunes, are similar enough to be classified under the same name. Due to its heavy use by pioneers in the field, they chose the generic term, dune, to describe these features. A further distinction was made with the modifier, "subaqueous", in order to distinguish underwater dunes from their aeolian counterparts and a "small", "medium", "large", or "very large" designation was applied. This classification scheme did not find universal acceptance, and in its place, the ASCE Task Force's system remains as one of the most commonly adopted classification systems used in the literature. The ASCE classification, which is used in this report, also agrees with that given by Blatt et al. (1980) and shown earlier in Figure 8.

The linking of flow regimes and a sequence of bed configurations was made only after extensive flume investigations were conducted. Though no distinction was made between sand waves and dunes during early laboratory tests, the differences between these two types of bed forms were only made apparent after subsequent field and laboratory tests (Harms 1975). The flow regime concept has been successfully extended to riverine and tidal flows of large scale. Simons and Richardson (1963) identified the lower flow regime in alluvial channels by the existence of ripples and sand waves, bed forms that produce very little disturbance of the surface waters above or, in the case of dunes, water waves that are out of phase with the bed forms. In the upper flow regime, antidunes are in phase with water waves. The transition zone, which separates the lower and upper flow regimes, is a plane, flat bed, that is generally included in the upper flow regime to make it distinguishable from the lower-regime plane bed that evolves when ripples composed of sands coarser than 0.7 mm undergo flows too weak to produce sand waves (Blatt 1980). Sand ripples and waves characteristically differ in size, with ripples being smaller and not depth dependent if the water depth exceeds approximately 10 cm.

5.3 Lower Flow Regime Bed Forms

The lower flow regime is comprised primarily of sand ripples and sand waves. The wavelength of sand ripples and waves has been found to vary more than wave height, so Stride (1982) advocates this parameter to be more useful for classification purposes. For quartz-density sands tested under laboratory temperatures, the maximum wavelength of sand ripples is approximately 60 cm, but low height sand waves are usually several times larger than this. For sand waves with wave heights greater than the ASCE classification's cutoff of 5 cm between sand ripples and sand waves, wave lengths are several times larger than 50 cm; larger wave heights and wave lengths would appear for lower density sediments.

Flume studies conducted with quartz-density sand at 17°C reveal that there is a maximum median diameter above which ripples do not form as was distinguished in the ASCE Task Force's classification of transverse and plane bed states (Figure 12). Southard and Boguchwal (1973) determined that the maximum diameter under laboratory conditions is 0.7 mm, so the maximum median grain size increases for sediments of lower density and lower temperatures. Once the maximum grain size for ripples is exceeded, the initial plane bed, that would have formed ripples if smaller grains had been available, will be supplanted under higher currents and bottom stresses, with "very low-height sand waves" (Middleton 1977). Though the sand waves are less than 5 cm wave height minimum put forth by the ASCE Task Force, their wavelengths are greater than that of sand ripples; thus the "very low-height sand wave" designation. This stage is followed by sand waves that have longer wavelengths and wave heights that exceed 5 cm (Stride 1982).

Ripple/sand wave-instabilities may be assessed by predicting a curve of the minimum flow required to move a particular grain size; this determination also enables data points to be classified as either sand ripples or sand waves. Yalin's (1972) theory of dimensions states that sand ripples, formed under equilibrium, and sand wave data plotted together should be separated by this type of boundary on a Shields diagram. This curve is usually plotted in terms of dimensionless bed shear stress and grain Reynold's number, a measure of the relative importance of viscous action. A graph of various flume data sets taken for water depths greater than 12 cm was assembled by Johnson et al. (1981). Figure 13 shows sand ripples and sand waves plotted with two approximate data sets obtained by Hill (1971) in order to assess the instabilities between these two bed states. The sand waves may or may not have had sand ripples superimposed on top of them; if they did, the bed state was designated as being composed of sand waves. The scattering of data points on the graph shows an overlap of points for sand ripples and sand waves. Since it may be easier to identify sand ripples than very-low height sand waves in flume tests, some of the "ripple" data points may have been misclassified and in actuality be sand waves. So the boundary between ripples and sand waves shown in Figure 13 is the lowest sand wave boundary that passes through all but one of the sand wave designation points and those data points below the boundary may be considered with fair certainty to be ripples (Stride 1982).

Though flume studies have not specifically focused on determining the minimum grain size necessary for sand waves formation, flume data show that sand waves do not seem to form under laboratory temperatures for quartz sediments with a mean grain size of less than 0.17 mm. For sand waves that exhibit cross stratification in well-sorted sandstones, however, median grain sizes of less than 0.15 mm may be indicative of sand wave formation in warm tropical waters (Stride 1982).

5.4 Transition Bed

The transition bed is characterized by rapid changes in the bed and water configuration that accompany relatively small changes in flow conditions. With increasing flow intensity, the lower flow regime bed forms change progressively into a transition bed characterized by sand waves with increasing wavelength and decreasing wave height. The bed can become flat for fine materials under certain flow conditions and further increases in the flow can produce symmetrical sand waves that generally evolve and disappear quickly, without growing in wave height. This regime is unstable because the transition from washed out sand waves to a flat or "plane" bed progresses into symmetric sand waves that can cause boils on the water surface, a flat water surface or one that exhibits standing waves (Garde 1977). Sand wave observations made in the Brahmaputra River in South Asia by Coleman (1969) may be classified as transitional bed forms because they are long, almost flat-topped features with long wavelengths that move down stream much more quickly than sand waves do. These features are distinctive in that they are unlike the lower-regime sand waves; they are also unlikely to appear on continental shelves because Froude numbers there, a measure of inertia to gravity forces, are less than 0.1 (Stride 1982).

5.5 Upper Flow Regime Bed

In theory, the upper flow regime state only exists in environments with fine to medium sand and in shallow continental shelf waters less than 1 m deep. Flume studies show that when upper flow regimes with fine and medium sands experience increasing flow rates, the relatively planar bed evolves into antidunes. These features, sinusoidal to trochoidal in shape, are extremely temporal, move much more slowly than the surrounding flow, and are in phase with water waves Figure 8. Sediment continues to be transported down current with the individual grains being incorporated or pushed over the top of the antidune.

Distinctions are drawn between sand ripples and sand waves because of size, water depth constraints, and their origins. Antidunes require Froude numbers over 0.8 and very shallow water depths where currents are strong. Though they may occur atop tidal current sand banks or in intertidal areas, antidunes that form on the continental shelf are quite transitory in nature because of their small volume and high sand transport rates. Some symmetrical features in deep water depths and under the influence of weaker flows may have been incorrectly classified as antidunes when they are in fact wave-formed sand ripples or larger sand waves generated by uniformly flowing tidal currents (Stride 1982).

Bed phase stability does not lend itself to being accurately represented on a two dimensional diagram, though researchers' define the stability of bed configurations by representing bed phases on a size-velocity-depth diagram or a series of size-velocity or depth-velocity diagrams: Figure 14 shows bed phases for sand grains with a median diameter of 0.5 mm plotted on a graph of mean flow depth (cm) versus mean flow velocity (cm/sec); the mean flow depth shown in this figure is equivalent to the mean water depth. Figure 15 gives bed phases based on experimental data for flow depths of approximately 40 cm as a function of mean velocity (cm/sec) and mean grain size (mm). Since the water depths in the laboratory are not comparable to what is found in the ocean, however, use of such diagrams requires extrapolation of the boundaries found in a controlled environment to that which is found in the field. Consequently, researchers have sought to combine the depth and velocity parameters into one variable, either shear velocity or stream power, the latter being the product of mean velocity and shear stress. These allowances are successful, in part, because shear velocities in the field can be approximated in the laboratory by increasing the slope of the tank in order to increase flume velocities instead of increasing flume water depths.

6. SAND RIPPLES AND SAND WAVES

Their relatively low profiles have made sand ripples difficult to detect acoustically. Consequently, they have usually been identified by divers, underwater photography, or remotely operated television. Heavily dependent on near-bed flow, ripples artificially generated in a flume are therefore on the same magnitude as those found in the field. Yalin (1972) states that ripples can be distinguished from dunes by the following criteria: (1) average wavelengths of usually less than approximately 1600 times and greater than 600 times the median grain diameter of the sand from which they evolve, (2) wave height that does not exceed approximately 300 times the median grain diameter, and (3) both the wavelength and wave height are independent of depth if the water is deeper than approximately 10 cm. Sand ripple heights are approximately 1 to 5 cm, with wavelengths between 5 and 12 times the sand ripple height (Yalin 1972). As mentioned earlier, testing of quartz-density sands under laboratory temperatures at 17°C shows that ripples do not form when grain diameters exceed 0.7 mm (Southard 1973). Though the sands used in this experiment were well sorted, the finer grains in poorly sorted sands tend to form ripples, while the heavier grains migrate to the troughs.

Sand ripples with asymmetrical forms exhibit steep slopes on their leeward side and, for flows just above the threshold of grain movement, steep crests (Guy 1966). With increasing flow speed, sand waves are three dimensional and, as Allen (1968) recognized, the ripple crest generally remains transverse to the flow direction, though sand ripple shape can change with time for a particular position and even between different sand ripple groupings.

Ripples can be further broken down into three more descriptive categories based upon their primary forcing influence: asymmetrical current ripples, symmetrical sharp-crested ripples affected by the oscillatory water movements from sea waves and strong winds, and current-wave ripples. Hammond and Collins's (1979) flume experiment shows that the somewhat asymmetrical current-wave ripples have rounded crests and symmetry that is affected by current speed and oscillatory water speed. The wave height and wavelength of wave ripples composed of sand, on the other hand, increase rapidly with increasing grain size. For a gravel's mean grain diameter of 2.6 mm, Flemming and Stride (1967) determined that wave ripples can reach wave heights of 25 cm and wavelengths of 1.25 m. As was noted earlier in this report, various bed form classification schemes exist, each of which varies with respect to the definitions of their constituent bed forms and associated boundaries. This discrepancy is particularly evident when comparing Blatt's (1980) definition of ripples, which are said to be composed of silts to grains 0.7 mm in size, and Flemming and Stride's (1967) study in which it was determined that ripples could form with a mean grain size diameter of 2.6 mm.

Ripple formation progresses with the wavelength first increasing, leveling out, and possibly decreasing, with increasing water particle oscillation amplitude as determined by Mogridge and Kamphuis (1972). The effect of sea waves alone on sand ripples encourages the formation of two dimensional forms with crests that occur perpendicular to the advancing waves. Sea wave and current-induced sand ripples occur on continental shelves and anywhere that tidal currents sweep an area. They can also appear superimposed atop other bed forms but may disappear when strong enough waves or currents affect near bed flow velocities.

Larger in both wavelength and wave height than ripples, sand waves exhibit sinuous crests that are oriented oblique to the general flow direction. Flume-derived ratios of wavelength to wave height for sand waves are generally greater than 15 (Table 1). As mentioned in Section 3.2, a controversy exists over the terminology to use in describing sedimentary features. The term "dune" is used by hydraulics researchers to describe sand waves in flumes and rivers while marine geologists use "sand wave" and "megaripple" terms; the designation "megaripple" is used to describe those subtidal or intertidal bed forms with wavelengths between 0.6 m and 6 m in order to contrast with bed forms found in aeolian environments. The term sand wave has also been applied to all wave-like features composed of granular material, be it sand ripples, bars in rivers, and upper flow regime bed forms that are classified as antidunes by others. Stride (1982) considers the term, sand wave, to be applicable to any subaqueous transverse bed form that exists in the lower flow regime, is composed of sand,

and has wavelengths larger than sand ripples. In the case of two different sizes of sand waves occurring simultaneously, they can be distinguished from each other by further qualifying the term with the designation "small "or "large." The differences in sand wave dimensions and lee slope angle are all attributable to the turbulence in a flow.

7. SAND WAVE DIMENSIONS IN THE FIELD

Though flume investigations provide sand wave size data that are less variable than that found in the field, it is the many different combinations of flow depth and boundary conditions in nature that ultimately affect a bed form's morphology and distribution. It is known that bed form type, shape, dimensions, and orientation are quite variable over even the shortest of distances and no generally accepted theory exists for their origin. Nonetheless, the measurement of a sand wave's height, wavelength, grain size, and sometimes water depth can help to quantify the sediment transport at a particular site (Table 3).

7.1 Wave Heights

Sand wave heights are quite variable in the field. They have been identified in St. Andrew Bay, Florida by Salsman (1966) with heights between 0.6 m and 30 m; 30 m high sand waves on the Cultivator Shoal of Georges Bank were also surveyed by Jordon (1962). The limited water depths of estuaries do not preclude the existence of sand waves; Dalrymple (1984) surveyed sand wave heights of 0.8 m in the Bay of Fundy and Aliotta and Perillo (1987) identified heights of up to 6.0 m in the Bahia Blanca Estuary in Argentina. Boggs (1974) and Karl et al. (1986) state that the average sand wave height existing along continental margins, straits, and submarine canyons is 5 m. On the higher extreme, Harvey (1966) found average sand wave heights of 15 m in 150 m water depths in the St. Georges Channel of the Irish Sea (Levin 1992). Fenster et al. (1990) found large sand waves with wave heights up to 17 m in Eastern Long Island Sound; these features were found to be relatively stable over a 7 month period in that they migrated less than the horizontal accuracy limits of navigation (2 m).

Sand wave height can in some cases be influenced by prevailing meteorological events. For example, the ferocity of storm waves can act to diminish the size of sand waves, as is evidenced by the gale force winds affecting North Sea wand waves (Terwindt 1971); the quiescent period in between storms enables the bed forms to rebuild themselves to pre-storm heights. A protective layer of ice can also help to preserve sand waves; in the Gulf of St. Lawrence, Canada, distinctive sand waves are able to reach maximum heights and remain undisturbed below the ice where the water surface is forced to remain stable and air/sea interactions are minimized (Reinson 1979; Levin 1992).

7.2 Wavelengths

Although the classification of sand wavelength differs with various researchers, Boothroyd and Hubbard (1974) quantify the limits as ranging from 6 m to 1000 m, the upper limit identified on the Cultivator Shoal of Georges Bank (Jordan 1962). The shallow depths in estuaries or rivers can also exhibit a large variety of wavelength: sand wavelengths of 6 m have been identified in the Netherlands' Westerschelde Estuary by Terwindt and Brouwer

(1986) while wavelengths of up to 330 m were measured in the Ishikari River of Japan by Itakura et al. (1986).

7.3 Water Depth and Velocity

The water depths in which sand waves are found are as varied as their wave height and wavelength (Table 3). Water depths of less than 3 m have been known to contain sand waves with heights less than 0.5 m and wavelengths of 6 m (Terwind 1986). On the other extreme, sand waves have been surveyed in the Navarinsky Canyon head, Bering Sea in almost 500 m of water, a deep marine environment that is highly influenced by internal waves directed in a down-canyon direction (Karl 1986). In practice, however, the minimum water depth for sand wave formation is generally equal to 5 m. Of particular note is the fact that the depths in which sand waves are identified have no relationship to the water depth in which the sand wave was formed (Levin 1992). Water depth variations can also occur periodically and only in certain cases will they initiate sediment transport. Additionally, high river discharge rates from spring runoff, such as has been identified by Whetten and Fullen (1986) in Oregon's Columbia River, increase water depths which, in turn, promotes sand wave growth and migration. When water levels return to more normal levels, sand wave growth remains weak.

As noted earlier, the lee side of the sand wave points in the down current direction. Sand waves in rivers generally face downstream, but in the case of a saltwater wedge intrusion, the feature may migrate upstream. Flow speed is also a factor that influences bed form growth. Boothroyd and Hubbard (1974) noted sand wave migration in the Parker River Estuary for currents of 60 cm/sec; critical sand erosion velocities ranging between 0.5 m/sec and 0.6 m/sec were noted by several authors (McCave 1971; Terwind 1986; Whetten 1986). Note the slight variation in values with respect to the stream values given in Section 3-1, Figure 11, by Postma (1967).

Bed form size is greatly affected by flow power, water depth, sediment size, and shear velocity. Flow power is a function of both current speed and the slope of the water surface; shear stress is influenced by bed roughness. Zarillo (1982) uses the flow regime in Georgia's Duplin River, to demonstrate the affects of these parameters on bed form growth. Water retention in this tidal marsh area promotes the growth of a large surface water slope during the first half of ebb tide, which, in turn, causes high flow power and shear stress values during this stage of the tidal cycle. Since the incoming tides' flow and shear velocities are not strong enough to change the bed form shape, this reversing tidal current acutely affects bed form growth during ebb flows. Thus the maximum flow velocity, shear velocity, and flow power can be used to assess the occurrence and stability of a bed form.

Sand wave symmetry is affected by unidirectional currents and reversing tidal currents, the latter of which can be noticeably affected by spring and neap cycles (Stride 1982). Klein (1970) identified the affect of bottom current velocity asymmetries on sand wave orientation in the Minas Basin, Nova Scotia, where the mean sand grain size of sand waves is coarse and the threshold velocity for these sediments are greater than 65 cm/sec. He noted that sand waves whose lee side pointed in the ebb direction formed under maximum ebb tides of 90 cm/sec, while flood currents in the same channel never exceeded 65 cm/sec. For other channels in the basin, sand waves were oriented with their lee side pointing in the flood direction, since the flood tide of 90 cm/sec exceeded the ebb tide current of 65 cm/sec. Thus, maximum tidal flow, be it either ebb or flood, constrains sand wave migration and, thus, bed form orientation. Additionally, Harvey (1966) attributes the 15 m high sand waves in 150 m of water to current-induced density stratification in the St. Georges Channel area of the Irish Sea.

7.4 Grain Size

Mean grain size and sediment supply to an area also play a role in the formation of sand waves. As mentioned in the discussion of lower flow regimes (Section 3.3), there seems to be a minimum grain size necessary for sand waves to grow. Laboratory results show that quartz grains at 17°C do not form sand waves when their mean grain size is greater than 0.7 mm and the field results seem to indicate a correlation along these same lines; Table 3 relates sand wave formation and the availability of sufficient amounts of noncohesive soil in a particular size class (Levin 1992). Langhorne (1973) found that sand waves were absent from clay-filled areas of the Thames River Estuary. This fact may seem obvious, considering the nature of the word, "sand wave", but in some instances bottom soils are a mixture of different sediment types and grain sizes and not wholly uniform. In fact, dramatic changes in bed form type can occur at a site due to variations in the type of sediment supplied there; a portion of Florida's St. Andrew Bay, an area that had previously been devoid of sand waves but blanketed by silty clay, was affected by the construction of a jettied channel that ended longshore sediment transport of mud to the area. Resurveying showed that a new influx of sand was being introduced and sand waves were being formed in previously muddy areas (Salsman 1966).

Some authors have noted a distinct correlation between the amount of mud in bottom sediments and the lack of sand waves; see Table 3 for the grain size values of selected sand waves sites. Terwindt (1971) noted an absence of mud waves in bottom soils comprised of more than 15% mud and median grain sizes greater than coarse sand (0.5 mm). Bokuniewicz, Gordon, and Kastens (Bokuniewicz 1977) also drew a correlation between the lack of sand waves and the existence of bottom soils with more than 10% mud or more than 12% coarse sand. Increasing grain size also has been related to increasing sand wave dimensions (Zarillo 1982). Dalyrmple et al. (1978) and Zarillo (1982) determined that sand waves generally exhibit grain sizes between 0.25 mm to 0.5 mm.

7.5 Offshore Sand Wave Slope Angles

Stride (1982) states that for small asymmetrical sand waves reaching a maximum wave height of approximately 2 m, the lee slope angles are usually greater than 20°, but range between approximately 17° to 35°. On the other extreme, large asymmetrical tidal sand waves exhibit lee slope angles of approximately 2° to 30°, but usually less than 20°. Stoss slope angles are approximately equal to a lower value of 0.5° to 4° for strongly asymmetrical sand waves and approximately 14° for symmetrical waves. In fact, symmetrical waves may be thought of as being composed of steep slopes on both lee and stoss sides. Small sand waves of tidal origins generally exhibit steeper lee slopes than large tidal sand waves and steep avalanche faces are also more likely to form on small sand waves. Guy, Simons, and Richardson (1966) conducted flume tests on fine to very coarse sands and reported that the lee angle slopes of forms influenced by unidirectional currents was 20° to 32°, depending upon the amount of counter circulation from flow separation in the sand wave's trough. In general, the variability in slope values of ocean sand waves is due to the variation that exists between highly active forms with steep lee slopes and incomplete smoothed forms with gentle slopes.

8. SUMMARY AND CONCLUSION

As wavelike accumulations of transported sediments, subaqueous bed forms have been shown to be predominantly composed of consolidated, noncohesive sands. Their formation and transport is affected by a complex cycle of erosion and deposition that is affected by water flow rates, direction of flow, sediment supply, and individual sediment characteristics. The smaller features are temporal in nature and may last on the order of a few days or weeks, but the larger dunes and ridges can be built over hundreds of years by long-term bottom currents.

Though the morphodynamics of sand waves, and bed forms in general, is not known with certainty, the past few decades have seen scientists and engineers working with flumes to become better able to define the hydrodynamic conditions under which specific bed forms are stable. These many scientific investigations have focused on bed form morphology and general behavior in the laboratory, so the conditions of their occurrence seem fairly well established empirically in this environment, though extraplolation to field conditions is not directly correlatable. Nonetheless, much theoretical and field work remains to be performed on the origin and morphology of, for example, tidal current bed forms, because such relationships as growth mechanisms and the variability in morphology with respect to local conditions are yet to be fully understood.

Sand waves are worthy of additional study because the realization that sand waves may impact many engineering and navigational projects, such as shore protection and dredging of navigational channels, has required increasingly precise information about the shape and stability of the sea bed. Geophysical surveys of bed forms need to be made to provide information for improving our quantitative understanding and predictive capability. Research has shown that successful numerical simulation of sand waves is possible though detailed field surveys are required to calibrate the model.

Since the research on these complex and varied features remains far from complete, it is not possible to provide summary information regarding the probable distribution of sand waves throughout the world's oceans. In order to understand the mobility of sand waves at a particular site, however, it is imperative to pre-survey the area to quantify the extent of these features. It is suggested that for experiments short in duration, approximately one year or less, the absence of sand waves in the pre-survey may indicate that these features will not appear over the duration of the experiment. For areas where permanent installations will be placed, it would be beneficial to conduct such basic research as developing a numerical model using long-term statistics for the extreme environmental conditions that may form sand waves. Overall, researchers still strive toward a complete understanding and model of the full problem; these scientists and engineers will continue to explore and offer a number of explanations for bed form shape, development, patterns of occurrence, stability, and migration rates in order to better understand both modern sedimentary features and those in the stratigraphic record.

9. FIGURES AND TABLES

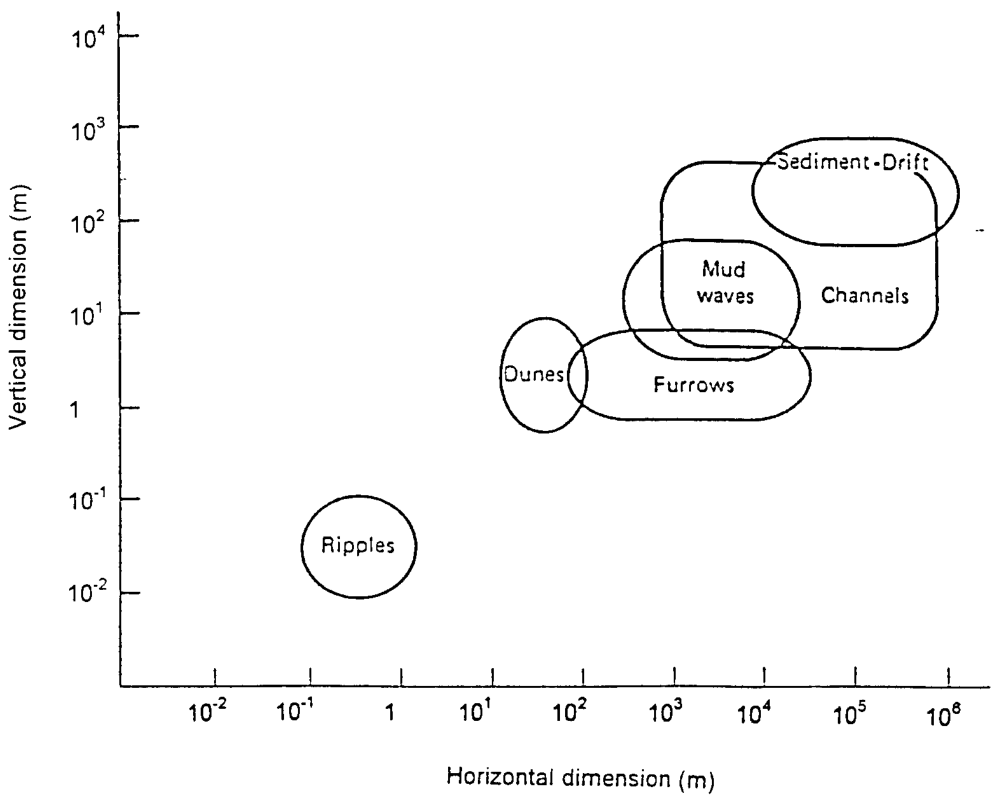


Figure 1. Horizontal and vertical dimensions of ocean bottom features (after Kennett 1982).

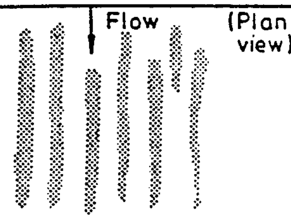
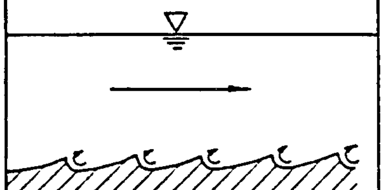
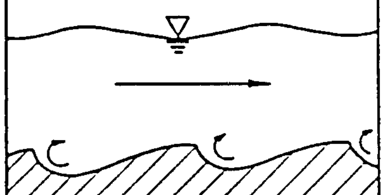
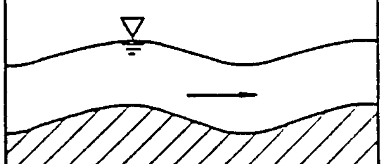
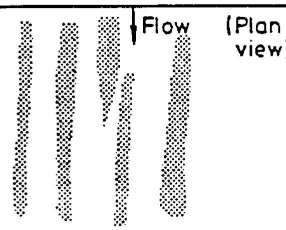
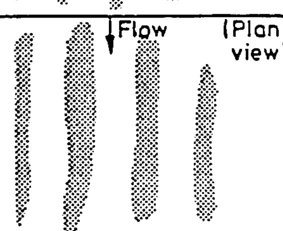
Bed form	General outline	Typical height
Parting lineation		1-5D
Ripples		10-100 D
Dunes		100-10000 D
Antidunes		
Sand ribbons		50-500 D
Tidal current ridges		7-30m

Figure 2. Steady flow bed forms (after Sleath 1984).
Bed form height is a function of median grain size of the sediment, D.

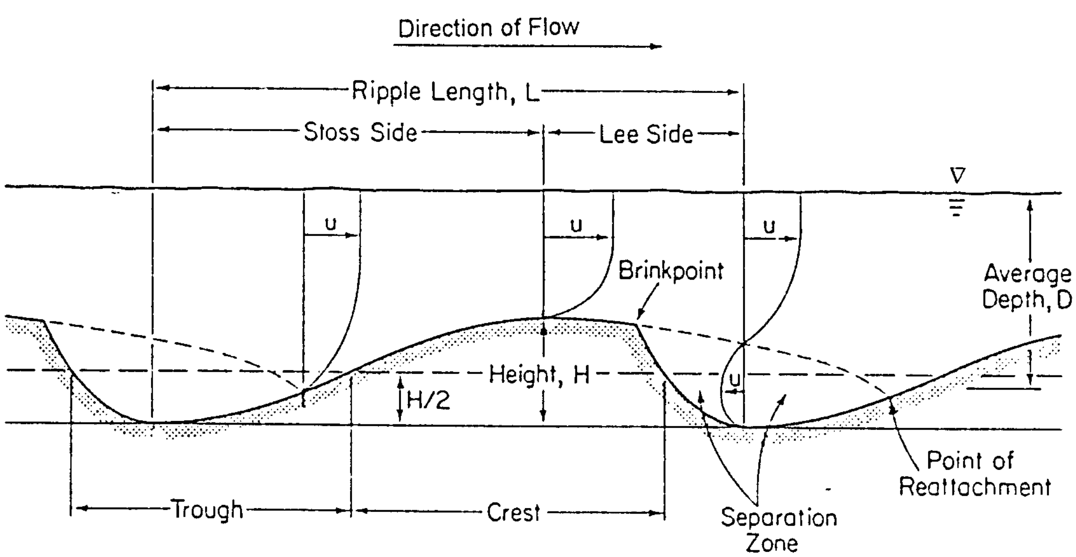


Figure 3. Cross-section of a bed form (after Allen 1968).

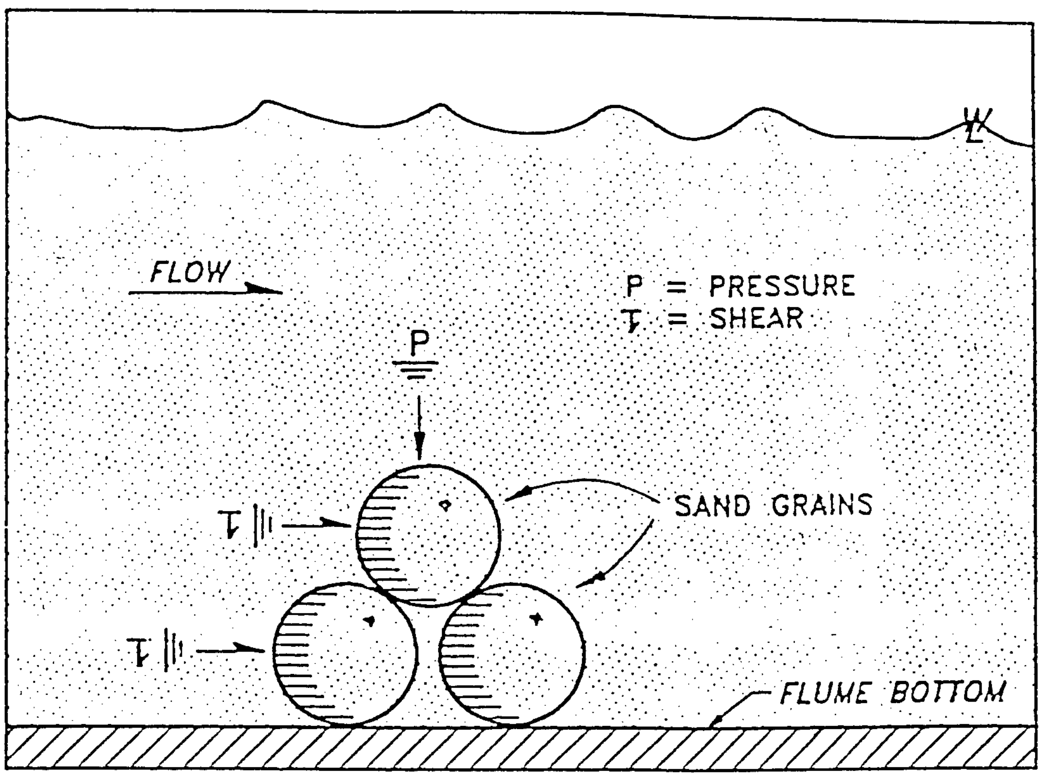


Figure 4. Water forces on grains (Levin 1992).

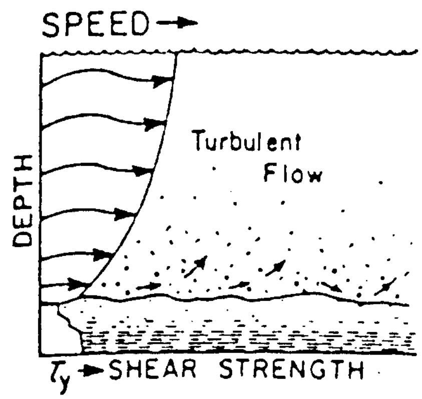


Figure 5. Velocity profile shows the reduction of current speed with depth in a turbulent flow field (After Nichols 1985).

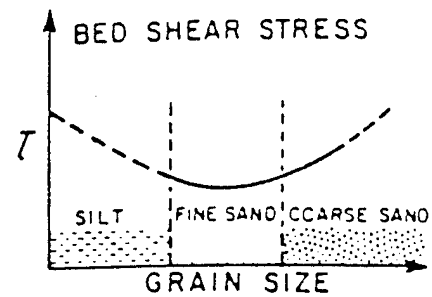


Figure 6. The variation of shear stress that is exerted on the bed is shown as a function of grain size (After Nichols 1985).

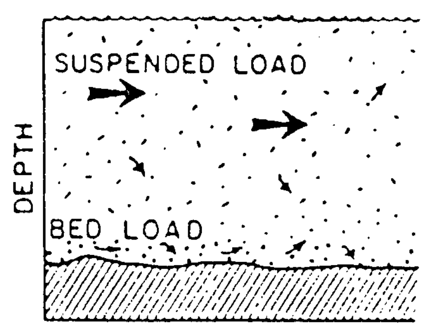


Figure 7. Distribution of suspended load in relation to bed load (After Nichols 1985).

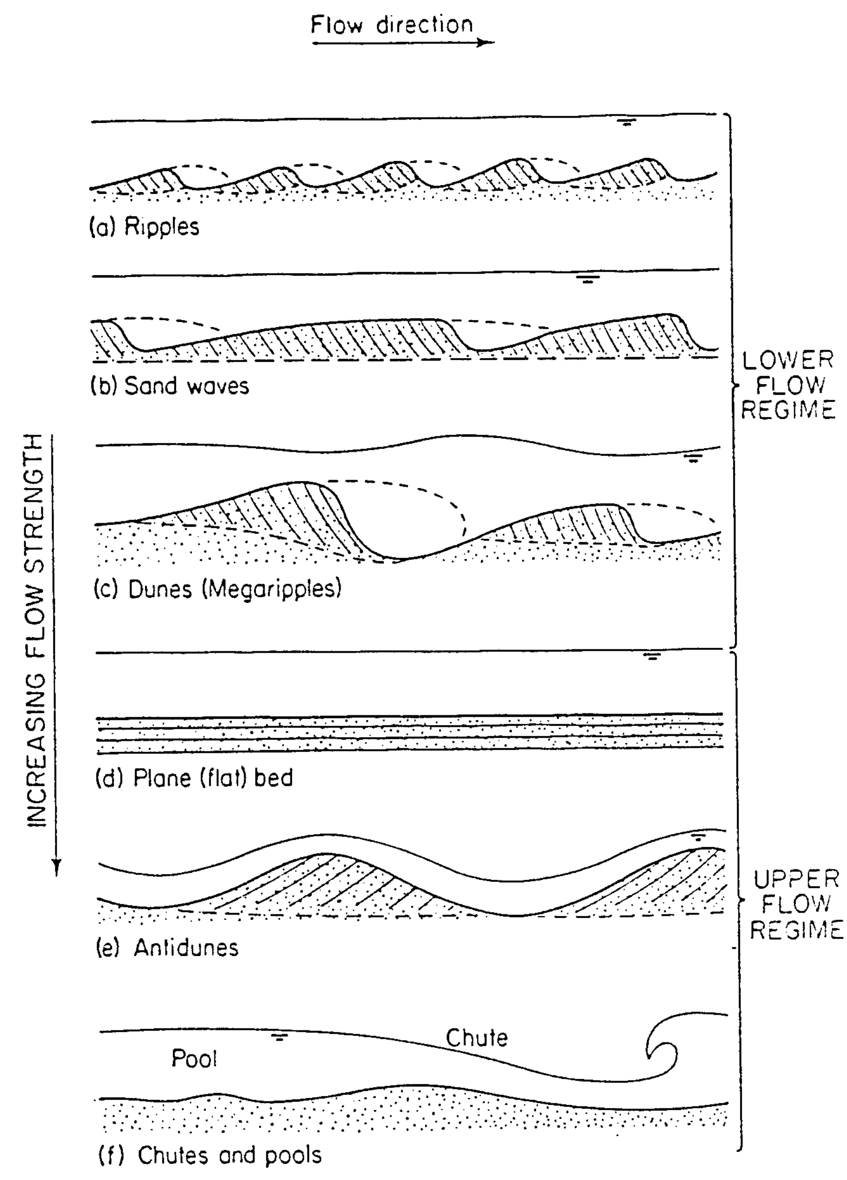


Figure 8. Types of bed forms in quasi-equilibrium unidirectional flows (After Blatt 1980).

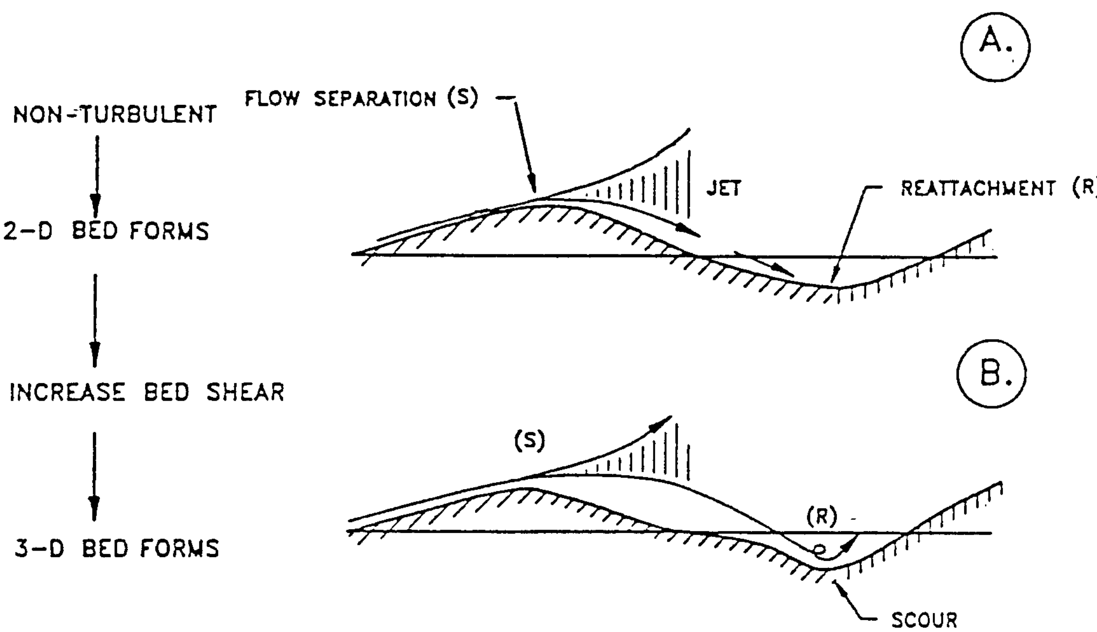


Figure 9. Jet current behavior over a ripple (After Levin 1992).

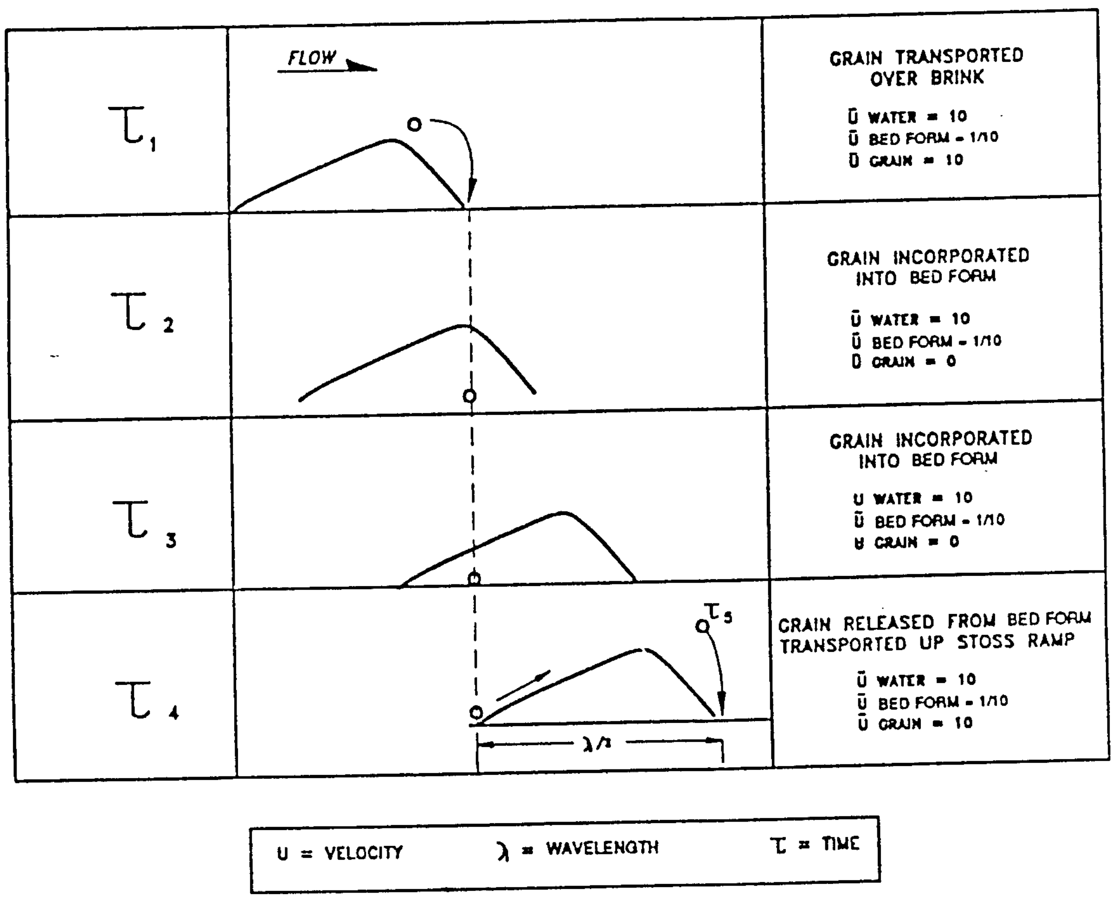


Figure 10. Bed form migration with respect to a single grain (After Levin 1992).

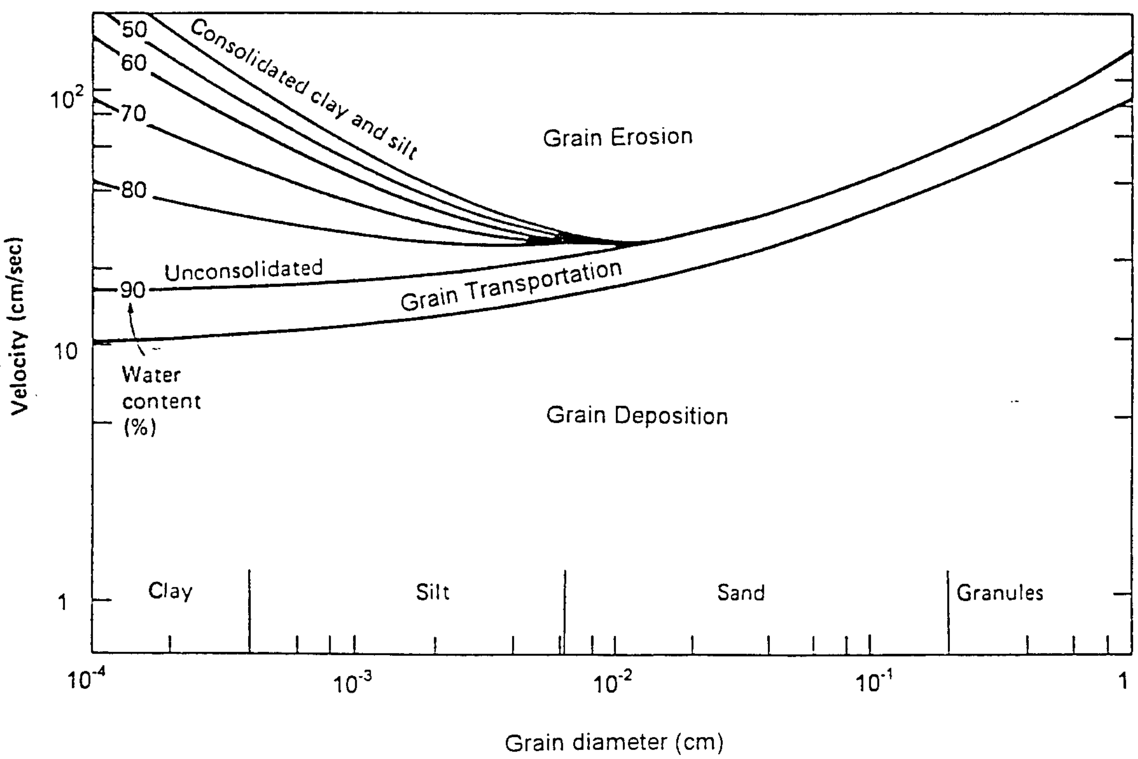


Figure 11. Bottom-current velocities for erosion, transportation, and deposition of sediments according to grain size (Postma 1967).

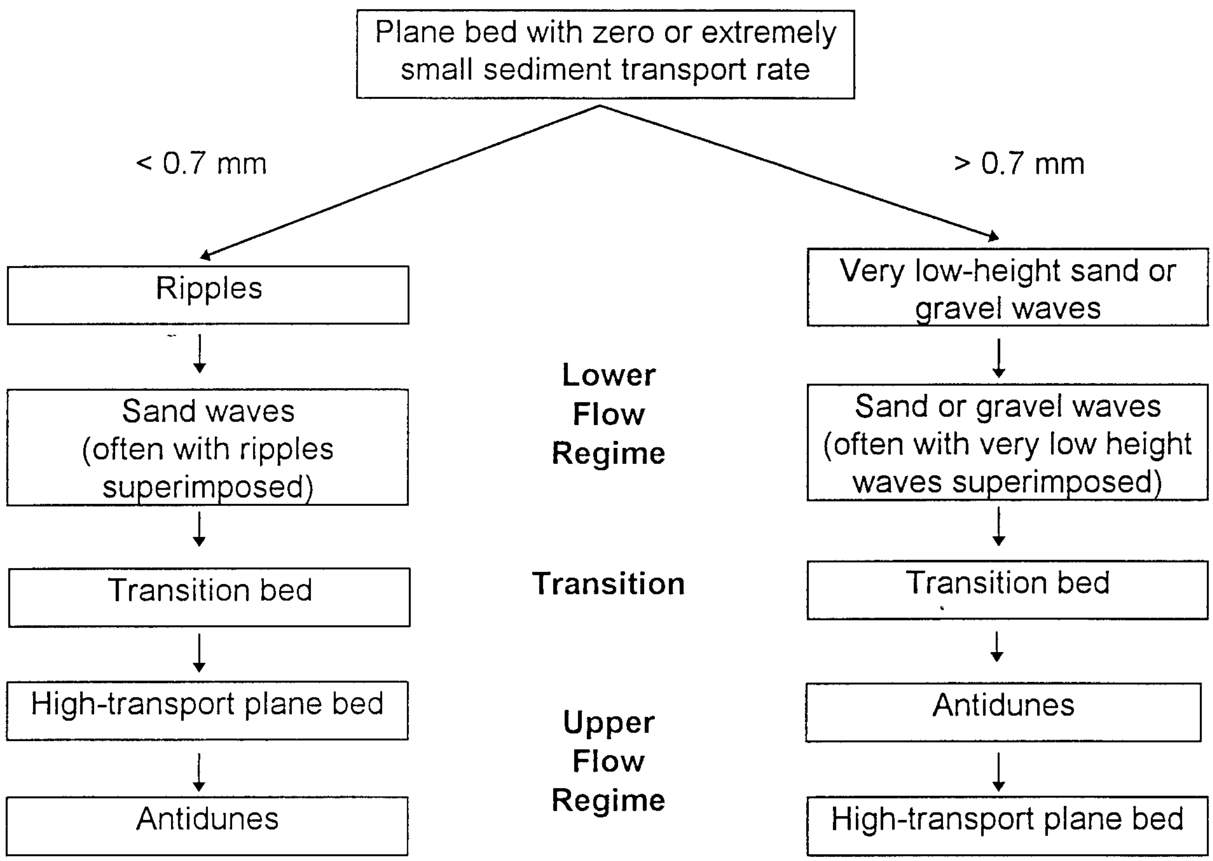


Figure 12. ASCE Task Force's classification of transverse and plane bed states in flumes using data for quartz density sand at about 17°C for median diameter less than and greater than 0.7 mm (A.S.C.E. 1966).

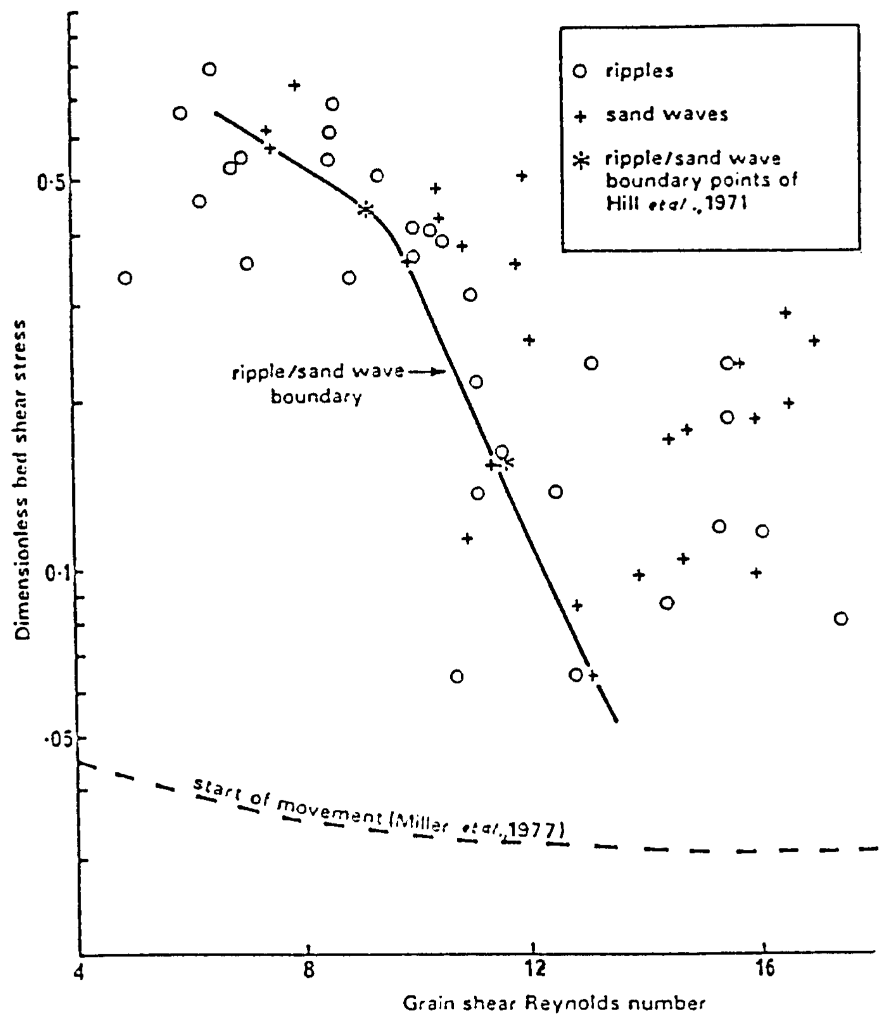


Figure 13. Reynolds Number for Selected Flume Data Sets (After Johnson et al. (1981).

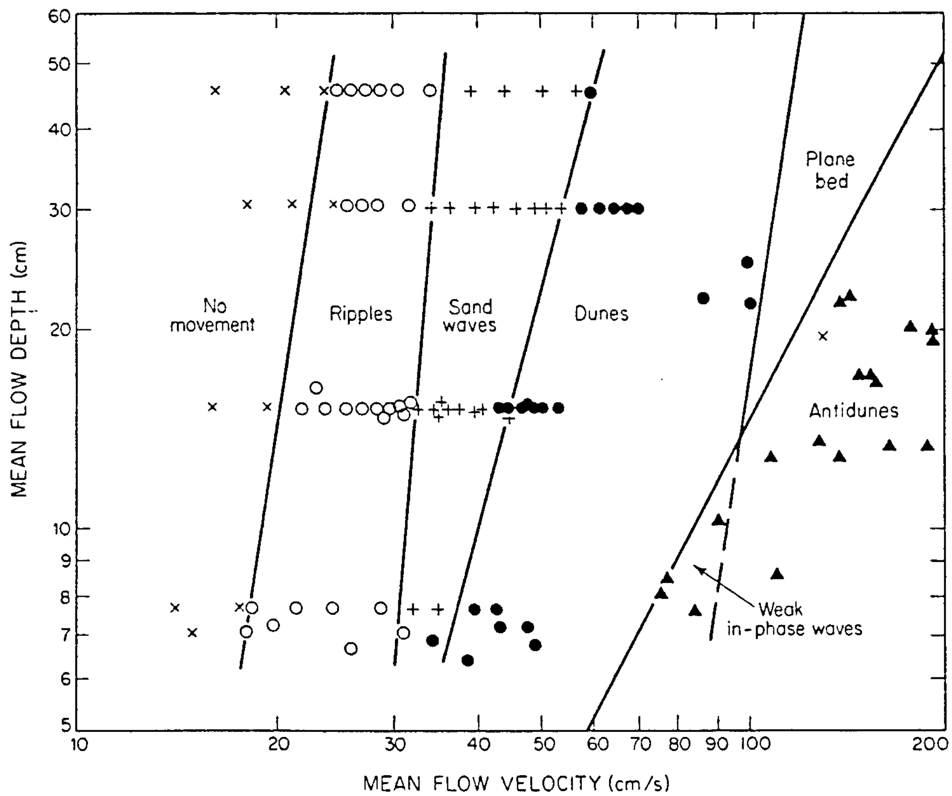


Figure 14. Depth-velocity diagram showing bed phases based on experiments with sands whose median size is approximately 0.5 mm (After Harms 1975).

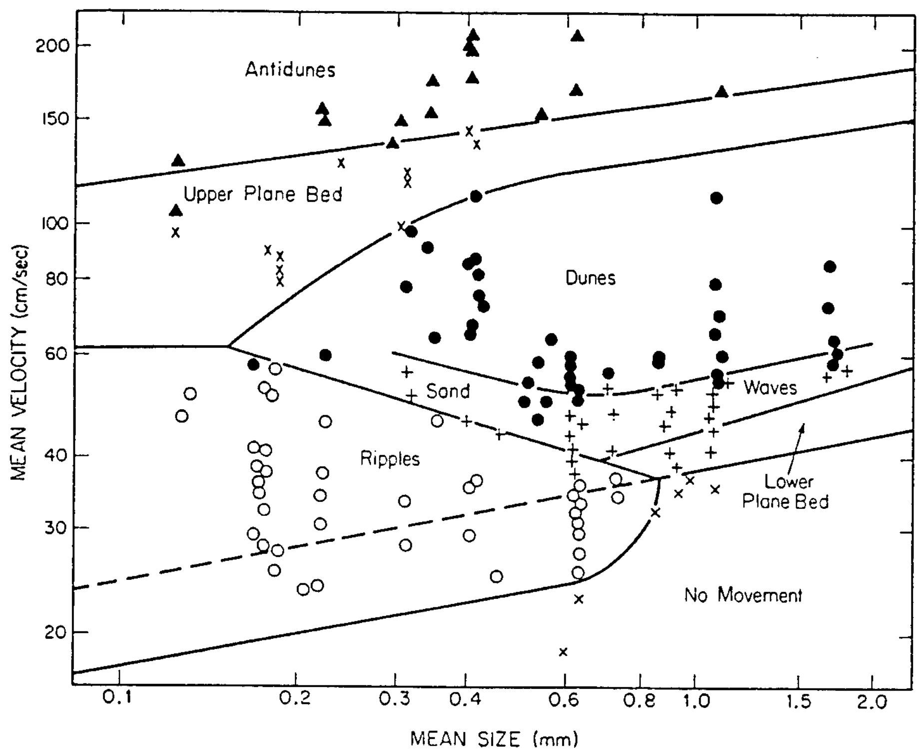


Figure 15. Velocity-size diagram showing bed phases based on experiments with flow depths of approximately 40 cm (After Middleton 1977).

Table 1. Characteristics of Bed Forms (After Blatt 1980).

	Ripples	*Sand waves*	*Dunes*	*Antidunes*
Length (chord)	0.1 to 0.3 m	5 to 100+ m	0.5 to 10 m	up to 5 m
Vertical Form (ripple) Index (chord length)*	8 (coarse sand) to 20 (fine sand)	15 to 100+; generally 50. Spacing may be variable	5 to 50 (larger values for coarser sizes)	7 (minimum) to 100+ generally high
Geometry	Straight to sinuous crested when first formed; otherwise generally linguoid	Sinuous crests, may be oblique to flow, no deeply scoured troughs	Sinuous to lunate, deeply scoured troughs	Sinusoidal to trochoidal or 3d, symmetrical, regular forms; lee face at less than angle of repose
Grain size	silt to 0.7 mm	Coarser than 0.25 mm	Coarser than 0.1 mm	? no limit ?
Controlling hydraulic variables	Shear velocity (or velocity and depth)	Shear velocity (or velocity and depth)	Shear velocity (or velocity and depth)	Froude number and depth: minimum Fr=0.8 for shallow depths.

* Vertical Form Index (wavelength-to-wave height ratio)

Table 2. Duplication and overlaps in bed form nomenclature (A.S.C.E. 1966).

1.	Bed Configuration:	bed geometry forms of bed roughness bed form bed regime bed phase bed irregularities sand waves bed material forms bed shape
2.	Flat Bed:	smooth bed plane bed
3.	Bed form:	bed irregularity bed wave bed feature dune ripple sand bar gravel bar sand wave
4.	Ripples:	dunes sand waves ripple marks current ripples
5.	Bars:	sand waves banks sand banks deltas slipoff points
6.	Dunes:	ripples sand waves sand bars
7.	Transition:	sand waves washed-out dunes
8.	Antidunes:	standing waves antiripples sand waves
9.	Chutes and Pools:	violent antidunes

Table 3. Sand wave dimensions around the world (After Levin 1992).

Location	z*	H	L	Sand	Vel
Columbia River, Oregon	10	5.5	200	--	0.6
Thames Estuary, G.B.	20	8	--	0.14	--
Sapelo Island, GA	6	0.8-1.3	15-45	0.3	1.0
Parker River Estuary, MA	6	0.5	>6	--	0.6-0.8
Westerschelde, Holland	<3	0.2	>6	--	0.5-0.6
St. Andrew Bay, FL	11	0.6	13-20	--	1.2
Long Island Sound	10	to 4	10-100	--	0.35-0.4
Weser River, Germany	4	to 3	--	0.5	1.0
Ishikari River, Japan	8	2.0	330	0.26	--
Bay of Fundy, Canada	17	0.15-3.4	13-40	0.13-1.6	0.5-2.0
Minas Bay, Canada	6.4	0.8	38	0.27-1.27	1.0
Gulf of St. Lawrence	3-5	2.0	60	--	--
Bahia Blanca, Argentina	10	6	80-200	0.2	--
Isle of Man, Irish Sea	11	6	125	--	0.3
Irish Sea	80-100	15	200	0.6-0.9	1.0
Navarinsky Canyon, Bering Sea	>175	5	650	--	--
SE African Continental Margin	50	0.8-8	7-200	0.6	0.6
North Sea	20	3-15	144-1,200	0.5-0.8	0.5-0.8

z* = depth in meters, depth may vary tidally
H = bed form waveheight, meters
L = bed form wavelength, meters
Sand = mean grain size, or range of sand size, mm
Vel = mean velocity, or range of velocity in sandwave field, m/sec
-- = no data available

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